Late Pleniglacial and Late Glacial aeolian phases in The Netherlands

Kasse, C.

published in
In: W. Schirmer (ed.) Dunes and fossil soils. GeoArchaeoRhein 3
1999

document version
Publisher's PDF, also known as Version of record

SONDERDRUCK AUS:

Wolfgang Schirmer (Editor)

Dunes and fossil soils

with contributions by

Barbara Antczak-Górka	Marion Müller
Margot Böse	Bolesław Nowaczyk
Arthur Brande	Iwona Okuniewska-Nowaczyk
Renata Dulias	Jolanta Pełka-Gościniak
Michael Facklam	Wolfgang Schirmer
Katarzyna Issmer	Norbert Schlaak
Klaus-Dieter Jäger	Tadeusz Szczypek
Cornelis Kasse	Jerzy Wach
Dietrich Kopp	Józef Wojtanowicz
Alojzy Kowalkowski	Steffen Wolters

LIT

GeoArchaeoRhein, **3**: 61–82; Münster 1999

Late Pleniglacial and Late Glacial aeolian phases in The Netherlands

CORNELIS KASSE

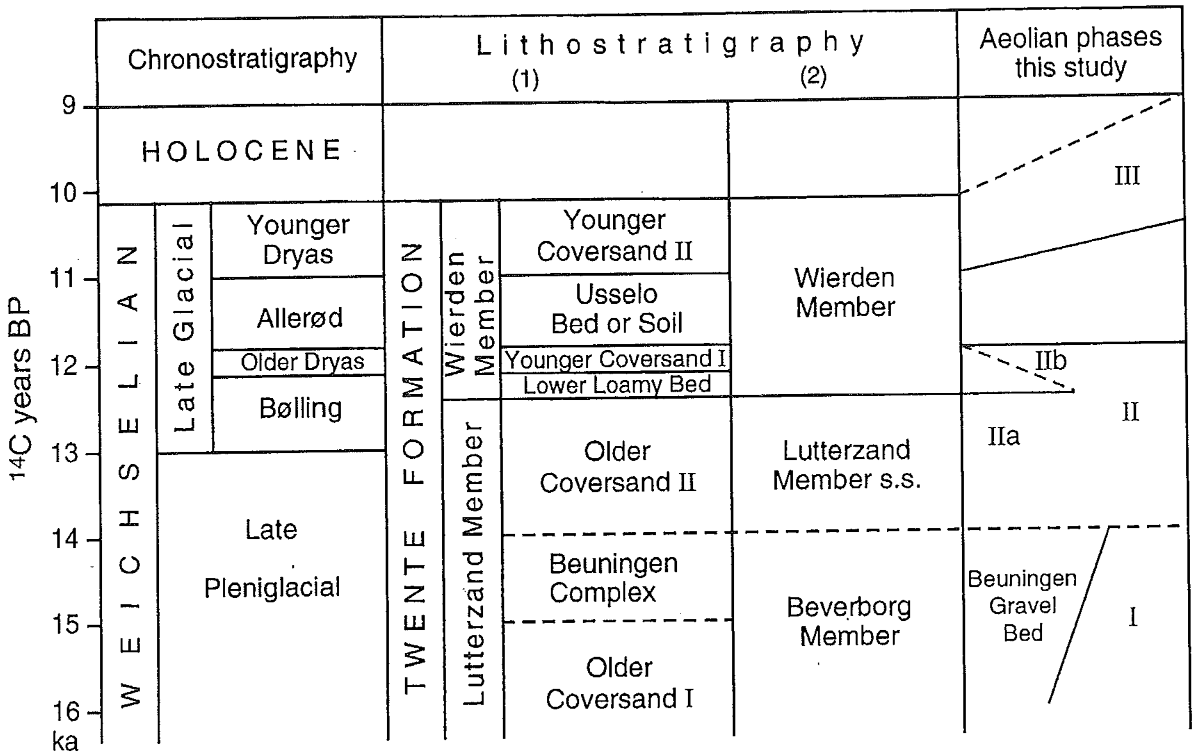
Abstract: A new Late Glacial site Oud-Lutten with well-developed aeolian units separated by organic deposits is presented from the eastern Netherlands. The results are discussed and a comparison is made with previously established phases of aeolian activity. Three aeolian phases are distinguished during the Weichselian Late Pleniglacial and Late Glacial in the Netherlands (see Tab. 1). During phase I (c. 20–14 ka BP) aeolian activity was very important, but the preservation of primary aeolian deposits was low due to the presence of permafrost and related overland flow. Phase II (c. 14–11.9 ka BP) represents the major period of widespread aeolian deposition in sand-sheets and low-dunes. Locally, two subphases (II a and II b) have been distinguished separated by Bølling organic sediments. It is stressed that subphase II b is not strictly coinciding with the Older Dryas chronozone. During phase III (c. 11–9 ka BP) localized aeolian deposition occurred mostly as dunes. Reworking of older aeolian deposits took place on the dryer interfluves and secondly, new dunes were formed on the eastern banks of rivers by deflation from river floodplains. The time of formation, geomorphology and sedimentary environments of the aeolian deposits are discussed and compared with similar sediments in central Europe.

1. Introduction

Aeolian coversands of the end of the last glacial are a very widespread deposit in western and central Europe. The occurrence of coversands is patchy in Great Britain (BATEMAN 1995) but on the continent the coversands form a continuous deposit in the Netherlands, Germany, Poland and Russia (KOSTER 1988; KASSE 1997; ZEEBERG 1998). The source areas were situated in the large delta plains of the Rhine and Maas and in the (pro)glacial sandy deposits along the Weichselian ice sheet limit. Coversand deposition took place especially at the end of the last glacial from the Last Glacial Maximum (c. 25 ka) onwards until the beginning of the Holocene.

VAN DER HAMMEN (1951, 1971) introduced a litho- and chronostratigraphic scheme of the Late Pleniglacial and Late Glacial aeolian deposits in the eastern Netherlands (Tab. 1). Four aeolian units were distinguished: the Older Coversand I and II and Younger Coversand I and II, which were separated by phases of erosion or soil formation (Beuningen Complex, Lower Loamy Bed, Usselo Bed or Soil). The aeolian units were rather rigidly connected with distinct glacial or stadial intervals of respectively the Late Pleniglacial, Oldest Dryas, Older Dryas and Younger Dryas. Later on this scheme was used also in Germany (DÜCKER & MAARLEVELD 1957) and Poland (e. g. KOZARSKI 1990;

Tab. 1: Chrono- and lithostratigraphy of the Late Pleniglacial and Late Glacial aeolian deposits in the Netherlands. (1) is according to VAN DER HAMMEN (1971) and VAN GEEL et al. (1989); (2) is according to VAN HUISSTEDEN (1990).



KOZARSKI & NOWACZYK 1991). Apart from stratigraphical research the sedimentology and depositional environments of the aeolian deposits have been investigated and more dates have become available (SCHWAN 1986). In this article the supposed climatic relationship between coversand deposition and cold phases will be evaluated and it will be demonstrated that the phases of coversand deposition are not so time constrained as has been suggested previously. For example the Younger Coversand II in the Netherlands has been deposited especially in the second half of the Younger Dryas period possibly extending into the early Holocene, although the first part of the Younger Dryas was colder. Besides coldness other environmental factors like aridity, preservation potential, regional vegetation differences and changes in fluvial systems have been important as well in the formation of coversands and dunes.

2. Results

A coversand profile has been investigated near the village of Oud-Lutten in the province Overijssel in the eastern Netherlands (Fig. 1). The profile consists of several layers of aeolian sand separated by organic deposits. The peaty deposits had been found during soil mapping of the area. The investigated site is located along the northern margin of the Vecht river valley: an east-west ice-marginal valley of Saalian age. Directly north of

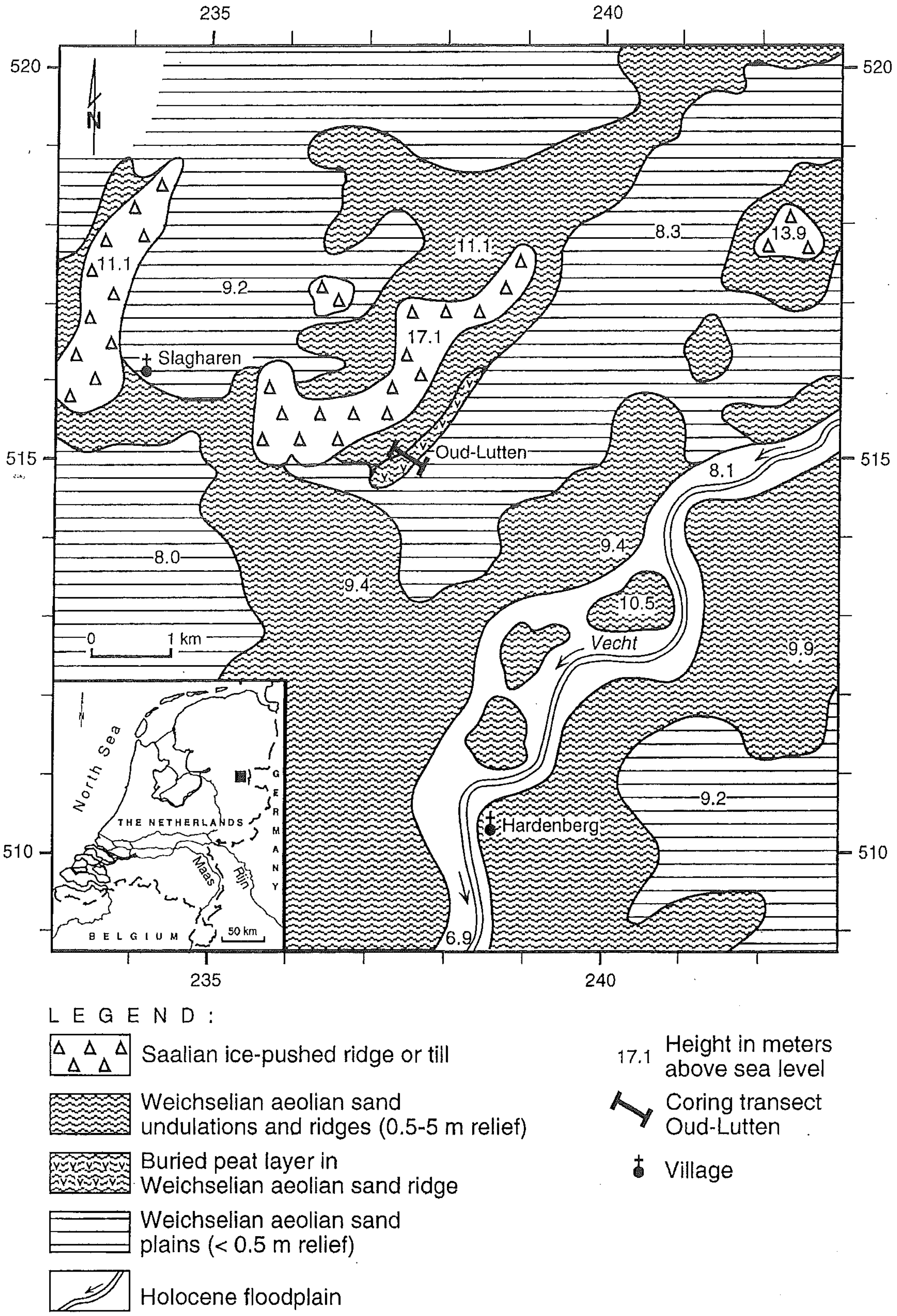


Fig. 1: Geomorphological map of the investigated area. The Oud-Lutten site is located in a coversand ridge at the boundary of the higher-lying coversand undulations to the lower-lying coversand plains.

Oud-Lutten the Saalian tills are found at the surface forming the higher landscape elements (Fig. 1). Following the Saalian, the Vecht valley south of Oud-Lutten has been filled with Eemian and Weichselian fluvial deposits c. 10–20 m thick. In the final arid stages of the Weichselian glacial the landscape was covered by wind-blown sands (so-called coversands in the Netherlands) and the Vecht river was confined to its present-day course. The coversands are up to several meters thick and especially around the ice-pushed ridges and along the Vecht floodplain the coversands are thicker and have more relief (Fig. 1). The investigated site Oud-Lutten is situated at the transition from the higher aeolian coversand relief in the northwest to the lower aeolian coversand plain in the southeast. According to the geomorphological map of the Netherlands (1:50,000) a coversand ridge is located at this transition.

Perpendicular to the coversand ridge a coring profile was made and afterwards a profile pit has been dug to sample the organic layers (Fig. 2). At the base medium to fine sand in a fining-upward sequence has been found which is interpreted as a braided river deposit of the Vecht during the Weichselian Late Pleniglacial (HUISINK 1998). This fluvial unit is overlain by a thin gravel bed, its components generally not exceeding 2 cm. This gravel bed is correlated with the Beuningen Gravel Bed which is a widespread marker horizon in the Netherlands, separating the so-called Older Coversand I and II. However, in this section the aeolian Older Coversand I is missing and the Beuningen Gravel Bed overlies fluvial deposits equivalent in age to the Older Coversand I (see also discussion). Overlying the Beuningen three aeolian units have been distinguished that are separated by organic deposits (Fig. 2). Aeolian unit A consists of fine-grained sand with loamy laminae. Especially below the organic layer the loamy laminae in unit A can be slightly organic containing some detritus and fine, possibly sedge roots. The upper boundary of unit A to the overlying organic layer is gradual; in the northwestern part of the section the upper boundary is indistinct. Aeolian units B and C are somewhat coarser grained than unit A because the loamy laminae are missing. Unit B is only present in the central part of the cross section; it forms an aeolian wedge intercalated in the organic layer. The aeolian sand grades laterally into a sandy moss peat and moss peat in a southeastern direction, indicating aeolian sand supply from the northwest. Unit B is time-equivalent with the lower part of the organic layer. The upper part of the organic layer is very pure without sand and is equivalent with the Usselo soil that overlies unit B. Aeolian unit C truncates the Usselo soil in the northwestern part of the cross section and overlies the organic layer with a sharp contact. The organic unit occurs in the central and southeastern part of the cross section and reveals a lateral and vertical differentiation in composition. It consists of peaty loam at the base with many fine sedge roots, changing upward into sandy peat with very well-preserved mosses. The upper part of the peat is more amorphous with some seeds of *Menyanthes*. Laterally, the central part is more sandy and moss-dominated, while the southeastern part is more loamy and contains less moss fragments and more sedge and *Menyanthes*.

The three aeolian units represent three fluxes of aeolian sedimentation at this site. The geomorphological position at the southeastern fringe of a coversand ridge and the cross

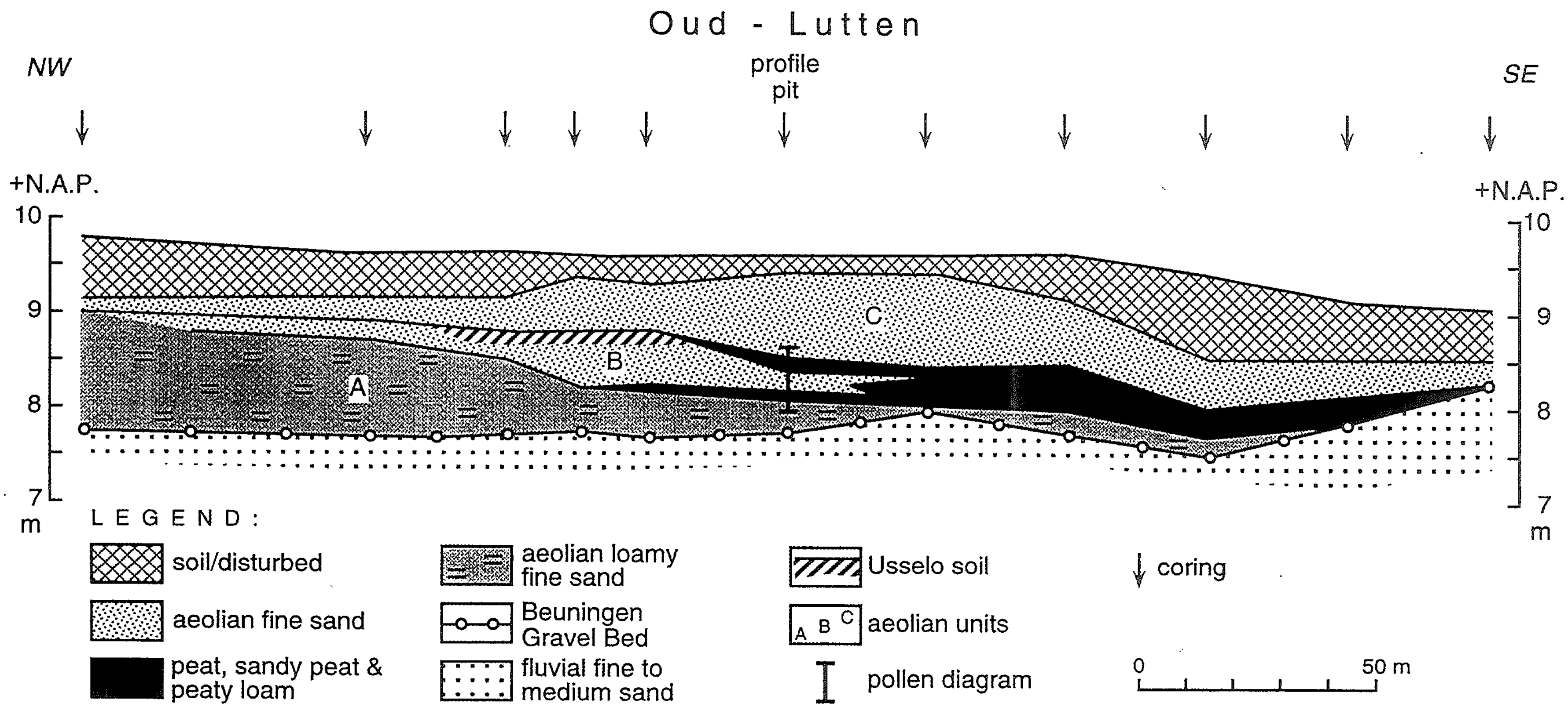


Fig. 2: Cross-section through the coversand ridge at Oud-Lutten showing three aeolian units separated by organic beds. For location see Fig. 1.

section both indicate that the described sequence has been formed by the step-wise migration of a low dune ridge to the southeast implying northwesterly winds during its formation. Because of this step-wise migration, organic deposits, that accumulated because of the wet seepage conditions at the lee side of the dune, were subsequently buried and therefore preserved only locally. To the northwest in the higher coversand area, standstill phases (e. g. soils) in the aeolian deposition have probably been eroded; to the southeast distinctive Late Glacial aeolian units have not been formed and Late Glacial organic material or soils have been incorporated in the Holocene soil profile. Only in the central part of the cross section three phases of aeolian activity are present and as a hypothesis it was proposed that the three aeolian units A, B and C might be correlated with the Older Coversand II, Younger Coversand I and Younger Coversand II respectively. In order to test this hypothesis a profile pit was excavated and two box cores (40 cm high, 10 cm wide, 10 cm deep) were taken for pollen analyses and radiocarbon dating (not available yet).

The pollen diagram shows seven pollen zones. The Cyperaceae are not included in the pollen sum. The biostratigraphic interpretations and tentative dates in conventional ^{14}C-years BP are based on HOEK (1997).

The lowermost zone 1a (102.5–137 cm) is dominated by herbs, especially Gramineae. *Pinus*, *Alnus* and *Picea* are high in the loamy lower part indicating reworking. *Artemisia* is rising towards the end of this zone, indicating the start of the Late Glacial around 12,900 BP. Therefore, this zone has been formed at the end of the Pleniglacial and beginning of the Late Glacial (Oldest Dryas). Aeolian unit A and the lower peat layer (Fig. 2) were all formed in this period.

Pollen zone 1b1 (93.5–102.5 cm) shows an increase of *Betula* and *Salix* while Gramineae is decreasing. This zone reflects the onset of the *Betula* phase of the Bølling dated at c. 12,450 BP.

Pollen zone 1b2 (70.5–93.5 cm) has lower Gramineae values and is dominated by *Betula* and *Salix*. The *Juniperus* curve is rising. This zone is interpreted as the *Betula* phase of the Bølling period between 12,450 and 12,100 BP. This means that the major part of the aeolian sand influx of unit B (Fig. 2) occurred during the Bølling in stead of the Older Dryas as was expected.

Pollen zone 1c (66.5–70.5 cm) shows a strong drop in the *Betula* values and Gramineae strongly increase. *Artemisia* attains high values. *Juniperus* attains its maximum values in this zone reflecting the stabilization of the aeolian landscape. This zone is equivalent with the Older Dryas zone (12,100–11,900 BP) (HOEK 1997). This means that only the upper part of sand unit B (Fig. 2) is of Older Dryas age.

Pollen zone 2a (63.5–66.5 cm) reveals high *Betula* values and low values of herbs. This zone is correlated with the *Betula* phase of the Allerød (11,900–11,250 BP). The aeolian sand of the previous zone has changed into a non-sandy peat indicating that the aeolian activity had ceased at the start of the Allerød.

Pollen zone 2b (57.5–63.5 cm) is dominated by *Pinus* and *Betula* has decreased. This zone reflects the *Pinus* phase of the Allerød (10,950–11,250 BP).

Pollen zone 3 (55–57.5 cm) shows a strong decrease of *Pinus* and increase of herbs. The vegetational change is in the uppermost part of the peat and marks the Allerød-Younger Dryas transition at c. 10,950 BP. Shortly afterwards the peat is covered by renewed aeolian activity (unit C in Fig. 2) during the Younger Dryas.

3. Discussion

Upper Pleniglacial (c. 25–14 ka)

The first evidence of strong aeolian activity in the Netherlands has been registered in deposits of the Last Glacial Maximum. In Oud-Lutten aeolian sediments from this phase are missing but they have been described from a nearby exposure in the Vecht valley (HUISINK 1998). In the Netherlands these deposits were formerly ascribed to the Older Coversand I (VAN DER HAMMEN 1971) and later to the Beverborg Member (VAN HUISSTEDEN 1990) (Tab. 1). Their lithostratigraphic position is below the Beuningen Gravel Bed which is a widespread erosional surface separating the Older Coversand I and II. Sedimentological studies revealed that these Older Coversand I sediments are generally not of primary aeolian origin (VAN HUISSTEDEN et al. 1986; VANDENBERGHE & VAN HUISSTEDEN 1988; SCHWAN & VANDENBERGHE 1991). They are generally found in river valleys where they form the top of a fluvial fining upward sequence (SCHWAN 1987; VAN DEN BERG & SCHWAN 1996; KASSE 1997). These aeolian sediments, therefore, have been reworked by shallow water and recently they have been described as fluvio-aeolian deposits (SCHWAN 1987; KASSE 1997). This means that the grain-size characteristics and the generally horizontal bedding resemble aeolian deposits but sedimentary structures like shallow, concave lenses of coarse sand, small-scale current ripple lamination and clayey/silty drapes indicate that the last depositional process was fluvial (KASSE et al. 1995). The rivers of this period resembled present-day High Arctic braided river systems with fluvio-aeolian sedimentation (GOOD & BRYANT 1985).

Similar deposits have been described from Germany (SCHWAN 1987; MOL 1997) and Poland. Especially in Poland the studies of quartz grain morphology clearly indicate that after 30 ka the content of wind-abraded grains strongly increases in the fluvial deposits of this period. Also here, primary aeolian deposits of this phase are very scarce or even absent (KOZARSKI 1990; MANIKOWSKA 1991; GOŹDZIK 1991).

This widespread reworking of the aeolian deposits in river valleys has been explained by the presence of continuous permafrost which could be deduced from the presence of diagnostic periglacial features like ice-wedge casts (KASSE 1997). The permafrost layer in the generally sandy subsoil will have reduced the infiltration of rain and snow-melt water and consequently overland flow will have been an important process at the surface (WOO & WINTER 1993). Under such conditions it is envisaged that aeolian sediments that had been deposited on interfluves and floodplains during winter were constantly removed by rain and meltwater in spring and summer. Consequently, despite strong aeolian activity, aeolian deposits were not frequently preserved. Especially on interfluves the Older Coversand I is often missing and only Older Coversand II deposits are frequently found overlying a gravel lag concentrate on older substratum (Tab. 1: phase I). The erosional products were transported by overland flow and shallow runoff towards the valleys and accumulated there on the floodplains. As a result, the fluvio-aeolian deposits of this period are mostly restricted to valleys, which contrasts with the following period when extensive aeolian deposits blanketed the whole landscape.

The age of the deposits is not well established. Organic remains are generally absent in the Netherlands (VAN HUISSTEDEN et al. 1986; KASSE et al. 1995) and in Poland (GOŹDZIK 1991; KOZARSKI 1993). Lithostratigraphic units with peaty beds underlying the fluvial and fluvio-aeolian deposits have generally been dated as Middle Pleniglacial (VAN HUISSTEDEN 1990; MOL 1997) and therefore the overlying fluvio-aeolian unit has been attributed to the Late Pleniglacial. The general occurrence of large ice-wedge casts in the fluvio-aeolian sediments has often been correlated with the maximal cold of the last glacial (c. 25–15 ka) which is in agreement with a Late Pleniglacial age. Luminescence dates are still rather scarce but the OSL-dates of 18 to 20 ka obtained from Germany confirm the Late Pleniglacial age (MOL & RHODES 1997).

The general absence of datable organic material can partly be explained by the extreme coldness of the High Arctic climate, but the general tendency in most fluvial sequences, with a decrease of fluvial deposition and increase of (reworked) aeolian sediment, seems to indicate that an increase of the climatic aridity occurred as well towards the end of the period (KASSE 1997).

The bedding types are in general dominated by weak horizontal bedding of fine sand often with an alternation of loamy laminae (VAN DER HAMMEN 1971; MOL et al. 1993; KASSE et al. 1995). In this respect these Older Coversand I deposits resemble the Older Coversand II deposits. However, in contrast to the Older Coversand II the Older Coversand I deposits contain shallow, concave lenses of coarse sand, small-scale current ripple lamination and clayey/silty drapes indicative for fluvial deposition. Furthermore, the Older Coversand I deposits are frequently disturbed by cryogenic processes associated with seasonal freezing and thawing and therefore the horizontal bedding has a wavy of crinkly appearance due to water saturation and water escape and sometimes the original horizontal bedding has been almost totally erased giving the sediment a massive structure (SCHWAN & VANDENBERGHE 1991). Vertical platy structures, possibly associated with intense surface cooling of the permafrost landscape, are a common periglacial feature in this unit (VINK & SEVINK 1971; MOL et al. 1993). The presence of continuous permafrost and its final degradation at the end of the period has resulted in the formation of ice-wedge casts and large-scale cryoturbations in the Older Coversand I deposits. By contrast such periglacial structures are missing in the Older Coversand II (KASSE 1997).

Late Pleniglacial and early Late Glacial (Oldest Dryas) (c. 14 ka–12.5 ka)

Aeolian sediments from this period have previously been called the Older Coversand II (VAN DER HAMMEN 1971) and later the Lutterzand Member s. s. (sensu stricto) (VAN

HUISSTEDEN 1990) (Tab. 1: phase IIa). Aeolian unit A in Oud-Lutten (Fig. 2) can be correlated with this Older Coversand II unit. The occurrence of the latter unit is widespread in the Netherlands as it is found on both interfluves and in valleys covering the pre-existing landscape and therefore it has been called coversand (KOSTER 1988). It is the most prominent phase of aeolian deposition and many of the younger Late Glacial and Holocene aeolian deposits have been reworked from this Older Coversand II unit. In the Netherlands this unit forms an aeolian sand sheet with in general low relief. In Oud-Lutten (Fig. 2: unit A) the relief is less than 1 m. Also in Poland, flat aeolian covers are dominant but small dunes up to 3.5 m have been described (GOŹDZIK 1991; MANIKOWSKA 1991, 1994) from this period.

The age of the Older Coversand II is not well known, due to the general absence of datable organic material while luminescence dating has been performed only sporadically. KOLSTRUP (1980) reported a date of c. 14 ka BP for the base of the Older Coversand II in the Netherlands and MANIKOWSKA (1991) dated some organic deposits at the top of a fluvial sequence underlying the coversands at c. 14.3 to 14.6 ka BP. The end of the Older Coversand II deposition is often correlated with the start of the Bølling period at 13 ka (see e. g. VANDENBERGHE 1991: Fig. 2), however, without any dating control. VAN GEEL et al. (1989) presented two AMS-dates of c. 12.8 and 12.9 ka BP from organic laminae within the Older Coversand II unit which indicates that deposition continued into the Late Glacial (Tab. 1: phase IIa). As will be discussed below the end of this widespread coversand phase can be extended until the start of the Allerød, since in many cases there is no clear boundary in the coversands under the Allerød soil. In such cases the Older Coversand II and Younger Coversand I form one unit because Bølling age sediments or organics are not present (Tab. 1: phase II). Only in a few cases Bølling age sediments have been reported in the coversand area of the Netherlands (see below) (e. g. VAN GEEL et al. 1989) and Poland (GOŹDZIK 1991: 58; MANIKOWSKA 1994: 120). Dates of c. 12.4 to 12.0 ka BP indicate that at that time aeolian activity had decreased but not stopped. In Oud-Lutten, the aeolian activity forming unit A decreased already before the *Betula* phase of the Bølling (see Fig. 3), i. e. somewhat earlier than 12,450 BP (HOEK 1997).

As has been stated in the introduction, it has been argued previously that the aeolian phases strictly coincide with glacial or stadial conditions (e. g. KOZARSKI & NOWACZYK 1991: Tab. 3). However, this does not hold true for the most widespread and prominent Older Coversand II deposits. In contrast to the deposits of the previous period before c. 14 ka BP which are characterized by ice-wedge casts indicating permafrost, the Older Coversand II sediments only contain thin frost cracks, cm-scale wide and up to 1 m deep. These frost cracks definitely point to less severe (= warmer) climatic conditions with deep seasonal frost, so it has been concluded that the most extensive coversand/sand sheet phase occurred during the very arid waning stage of the last glacial (KASSE 1997). The bedding types in the deposits of this period are of primary aeolian origin. Structures indicative for fluvial processes are generally absent, showing the strong aridity of this

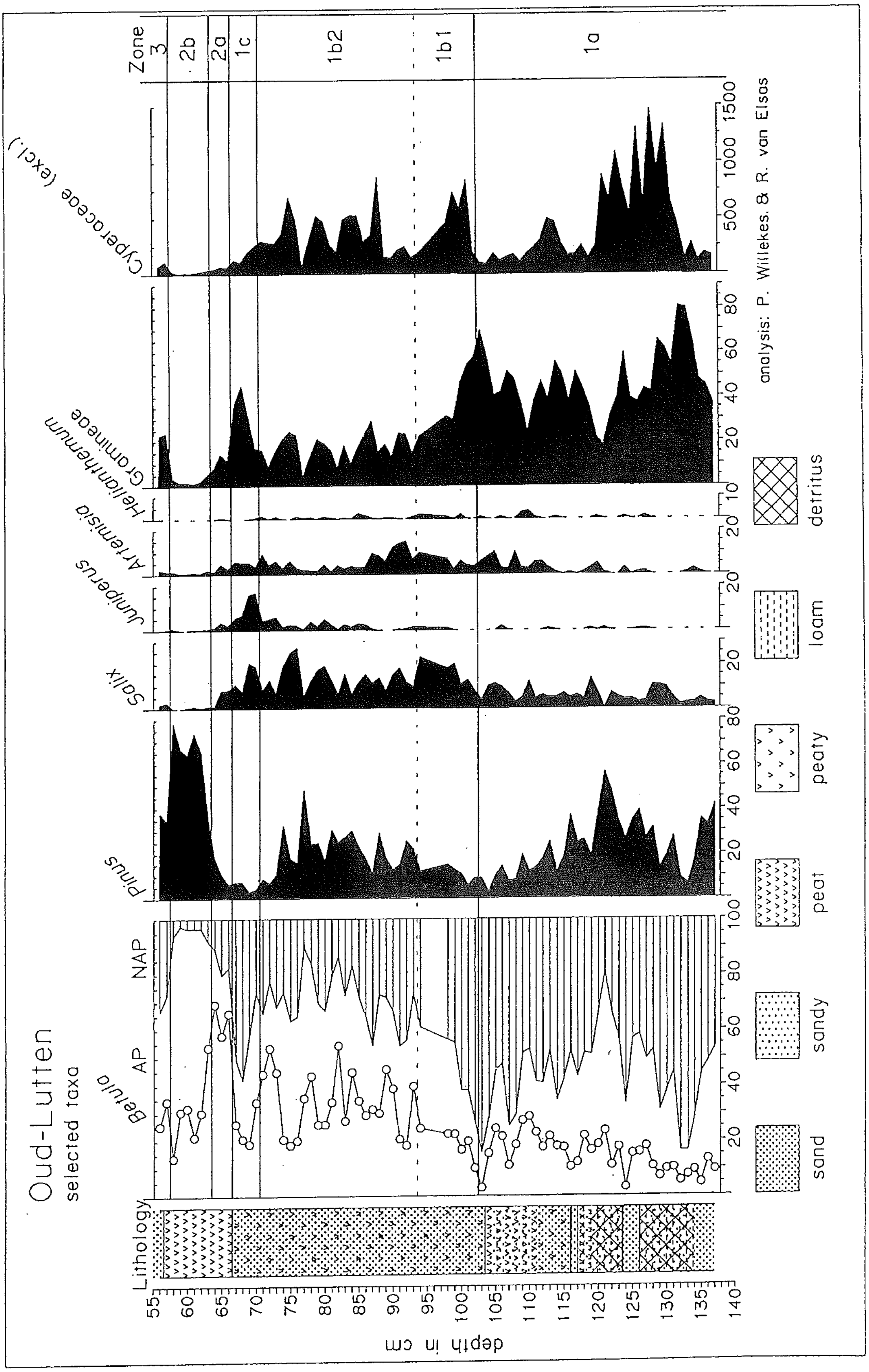


Fig. 3: Pollen diagram of the Late Glacial sequence at Oud-Lutten (selection of curves only). For location see Fig. 2.

period. Horizontal to low-angle bedding and alternating bedding of fine sand and silty fine sand are dominant in the sand-sheet deposits (KASSE 1997). The occurrence of silty fine sand or loamy laminae within the deposits has sometimes been attributed to permafrost conditions or niveo-aeolian deposition (STAPERT & VEENSTRA 1988). As has been stated above, however, evidence for the presence of permafrost has not been found. Niveo-aeolian deposits as observed in present-day arctic environments do not reveal loamy laminae and the disappearance of snow from the niveo-aeolian sediment results in sediment collapse and deformed bedding (KOSTER & DIJKMANS 1988; DIJKMANS 1990). The presence of loamy laminae in the coversands is presently explained by deposition and adherence of loessic material on a moist or wet surface (SCHWAN 1986). The coversands of lower topographic positions or poorly drained sites indeed have more frequent loamy laminae than on well-drained locations (VANDENBERGHE & KROOK 1981). The supposed strong aridity of this period seems to conflict with the widespread occurrence of loamy laminae that are indicative of surface wetness. It is postulated that despite the climatic aridity the depositional surface was wet in spring and early summer due to the combination of snow melt, thawing of deep seasonal frost and low relief of the sand sheet.

Bølling s. s. (c. 12.5–12.1 ka)

Bølling age organics or soils from the aeolian coversand landscape are extremely rare in the Netherlands, especially in comparison with the very frequent Allerød soils. The same holds true for Germany (SCHLAAK 1997) and Poland. MANIKOWSKA (1994) reported an initial soil from the Bølling period at Kamion dated at 12,235 BP. According to VAN DER HAMMEN (1971) the Lower Loamy Bed was formed during the Bølling s. l. (sensu lato) period, but this loamy layer has not been dated because organic carbon is missing. At Usselo VAN GEEL et al. (1989) reported organic deposits from the Bølling s. s. intercalated with Late Glacial aeolian sediments. More frequently Bølling age organics have been reported from pingo scars (DE GANS 1981; BOHNCKE et al. 1988).

The scarcity of Bølling age soils/organics in the aeolian district can in my opinion be explained by the continuation of aeolian deposition although with decreased intensity during that period. The land surface that was totally covered with aeolian sediments during the previous period (14–12.5 ka; Tab. 1: phase IIa) was only partly stabilized by vegetation and soil formation during the Bølling s. s. (12,5–12.0 ka). Evidence for this hypothesis is obtained from the few sites with Bølling organic deposits within the coversand region. In site Usselo the Bølling organic deposits are very sandy and contain sandy laminae which can be attributed to aeolian deposition at the same time of organic accumulation (KASSE 1997). In Oud-Lutten aeolian activity decreased and organic accumulation started before the *Betula* rise of the Bølling s. s. (see Fig. 3), this is shortly before 12,450 BP (HOEK 1997). However, already during the *Betula* zone of the Bølling (c. 12,450–12,100 BP) the aeolian deposition increased again (see Fig. 3) and most of the sand of unit B in Fig. 2 has been deposited during this *Betula* phase. This means that

aeolian deposition continued during the Bølling s. s. period although perhaps at a slower rate than before so that locally organic deposits could be formed (see Tab. 1: phase II). The start of the first organic accumulation in the Bølling seems to be around 12,500 BP (site Oud Lutten; VAN GEEL et al. 1989). This moment coincides with the Late Glacial rapid climatic amelioration at c. 14,670 cal yr BP as reconstructed from ice-core records (HOEK 1997: 48). The increased organic production may reflect a temperature and humidity increase associated with this climatic improvement. The development of a (sparse) vegetation cover can explain the change in this interval and the following Older Dryas from generally flat sand sheets to low dunes.
The formation of the Lower Loamy Bed (VAN DER HAMMEN 1971) intercalated in sandy coversands may also reflect the decreased aeolian activity, due to decreased sand supply and initial surface stabilization by vegetation during the Bølling. As the sand deposition became less important the deposition of loessic material — already existing as loamy laminae in the Older Coversand II — became more important.
As has been stated above, only in a few places the aeolian activity of the Late Pleniglacial and Late Glacial was interrupted by peat formation during the Bølling s. s. Only in such cases the aeolian activity can be separated in two phases (Tab. 1: phase II a and II b). Both examples (Usselo and Oud-Lutten) are situated at the lee side of a coversand ridge and it seems possible that peat formation during the Bølling was favored by water seepage in front of the coversand ridge. Except for these local peat accumulations, aeolian activity continued in most regions and the aeolian deposits can not be differentiated (Tab. 1: phase II). Aeolian activity finally stopped at the start of the Allerød.

Older Dryas (12.1–11.9 ka)

Previously the Younger Coversand I has been attributed rather strictly to the Older Dryas period (VAN DER HAMMEN 1971) (Tab. 1). This phase of aeolian deposition was later recorded as well in Germany and Poland (KOZARSKI & NOWACZYK 1991). However, in contrast to what was expected, the Oud-Lutten site shows that aeolian unit B (Fig. 2) was not restricted to the Older Dryas, since it has been formed in both the Bølling and Older Dryas periods (Tab. 1: phase II b). Furthermore, given the standard deviations of ^{14}C dates, it is questionable if an aeolian phase can be reliably correlated with such a short event like the Older Dryas.
According to VAN DER HAMMEN & WIJMSTRA (1971) the Younger Coversand I sediments generally do not contain loamy laminae like the preceding Older Coversand II and they are somewhat coarser grained. The bedding is in general horizontal to low-angle parallel, indicating that in the Netherlands the morphology was sand sheet and low dunes (KASSE 1997). The low dunes were formed as a last-stage feature during aeolian phase II, probably related to the vegetation development during the Bølling and Older Dryas periods.
VAN DER HAMMEN & WIJMSTRA (1971) sometimes subdivided the aeolian sequences underlying the Usselo soil on the basis of differences in bedding types, in a lower loamy-

laminated Older Coversand II part and an upper non-loamy horizontally laminated Younger Coversand I part (e. g. VAN DER HAMMEN & WIJMSTRA 1971: 96, 136). SCHWAN (1986), however, stressed that specific sedimentary facies need not be related to specific periods. The different sedimentary facies are the result of differences in topography, drainage and moisture conditions of the surface and these are to some extent time-independent. For instance, at the base of the Younger Dryas dunes deposits alternating bedding of loamy fine sand and sand — normally associated with the Older Coversand II — is present due to moister conditions at the dune base (SCHWAN 1991). So, bedding type should not be used as a time marker, but it has been noted that in any coversand sequence that formed during aeolian phase II (Tab. 1) alternating bedding of fine sand and loamy fine sand indicating a wet depositional surface, is always overlain by dry-aeolian horizontal to low-angle bedding without silty laminae (SCHWAN 1986: 104; KASSE 1997). This drying-up sequence in phase II seems to conflict with the assumed increase of climatic humidity at the start of the Bølling s. s. around 12.5 ka (see above). Despite the fact that the climate may have become wetter, the depositional surfaces may have become dryer due to the development of vegetation and associated low dunes during the Bølling and Older Dryas.

In contrast with the preceding Older Coversand II sands in the Netherlands, sediments dating from the Older Dryas are not so frequently reported, since so few sites exist in which both Bølling and Allerød organics underlie and overlie the Younger Coversand I. Mostly only the Usselo soil or peat of Allerød age is present overlying the coversand. Therefore, in the case of the Usselo soil overlying coversand, it is only allowed to say that the coversand predates the Allerød (Older Coversand II and/or Younger Coversand I) (Tab. 1: phase II).

In Poland on the other hand the Older Dryas period has often been ascribed as a period of major dune formation (KOZARSKI 1990; MANIKOWSKA 1994). Large-scale dune slipfaces covered by Allerød soils have been reported in several places and this seems to prove that the Older Dryas was a period of intense dune formation. However, as has been stated above for the Netherlands, also in Poland Bølling age sediments underlying the so-called Older Dryas dunes are rare (BOHNCKE et al. 1995). In many cases the dune sediments underlying the Allerød soil have been correlated with the Older Dryas period for morphological reasons without strong dating control (e. g. NOWACZYK 1986). It is the authors opinion that the difference in intensity of the Older Dryas aeolian phase in the Netherlands (low intensity) and Poland (high intensity) is only an apparent difference caused by differences in definition and insufficient dating. In Poland many dune sediments predating the Allerød soil have been assigned completely to the Older Dryas on morphological grounds, but the dune bodies could also have been formed over a longer time span covering the Bølling s. l. and the Older Dryas. In the Netherlands, aeolian sediments in the same stratigraphic position underlying the Allerød soil have often been assigned — because of the presence of loamy laminae — to the Older Coversand II of Late Pleniglacial age, but here also the aeolian deposition could have spanned the Late Pleniglacial, Bølling and Older Dryas.

The areal differentiation of sand-sheet and low-dune formation in the Netherlands and high-dune formation in Poland has been explained by climatic gradients (BÖSE 1991) and differences in vegetation development (KASSE 1997).

Allerød (c. 11.9–11.0 ka BP)

The Allerød period seems to be a period of non-deposition in the western and central European aeolian landscape (VAN DER HAMMEN 1951, 1971; VANDENBERGHE 1991; KOZARSKI & NOWACZYK 1991; MANIKOWSKA 1994; SCHLAAK 1997). The landscape was stabilized by a forest vegetation. In the Netherlands birch forests were succeeded by pine forest in the later part of the Allerød (c. 11,250 BP, HOEK 1997). In profile Oud-Lutten (Fig. 3) both the *Betula* and *Pinus* phase of the Allerød are present in the upper peat layer overlying aeolian unit B in Fig. 2. This means that aeolian activity has ceased before the *Betula* phase of the Allerød and the surroundings of the investigated site were stabilized by peat growth and soil formation (Fig. 2). Short-lived cold climatic oscillations during the Allerød like the Gerzensee oscillation did not affect the forest vegetation to such an extent that opening of the vegetation cover and deflation occurred.

The Usselo soil of Allerød age is a very common soil in Late Glacial aeolian sequences of the Netherlands. Human artefacts have been reported frequently from this soil horizon (HIJSZELER 1957). Its occurrence over the country as a whole, however, is not so widespread, because the soil has been preserved only when aeolian accumulation occurred during the Younger Dryas and the Younger Dryas aeolian deposition was rather local in comparison with especially the Older Coversand II deposition which was countrywide. In those regions where Younger Dryas deposition did not take place the Late Glacial soil has been incorporated by Holocene soil formation.

The appearance of the Allerød soil in the Netherlands can be very different. In low-lying areas in the aeolian region peat was formed (e. g. site Usselo, VAN GEEL et al. 1989; site Ossendrecht, SCHWAN 1991). In higher areas of the aeolian landscape a soil was formed (Usselo soil), 5 to 20 cm thick, that is normally characterized by a slight humus accumulation and some bleaching of the quartz grains (Ah-E-(Bw-)C profile). A weathering horizon (Bw) or weak illuviation of iron (Bs-horizon) have been found infrequently (site Weelde in Belgium, unpublished). Charcoal is a common constituent of the soil. ^{14}C-dates of 23 samples of charcoal from the Usselo soil cluster around 11,000 BP (HOEK 1997: 43). This agrees very well with the results from Poland, since MANIKOWSKA (1994: 121) reported that most of the dates of the Allerød soil are between 11,200 and 10,800 BP. The Usselo soil is intensively homogenized and the original sedimentary structures have been destroyed by bioturbation and soil forming processes. Round to oval-shaped burrows, c. 1 cm in diameter, are very frequent below the well-drained soils and they have commonly been attributed to burrowing activity by beetles (BRUSSAARD & RUNIA 1984).

In Poland the Allerød soil has been reported on the lee side of dune slip faces that were active during the Older Dryas (NOWACZYK 1986; MANIKOWSKA 1991). Recently, also in

Germany a soil of Allerød age (the Finow soil) has been found in Late Glacial dune sequences (SCHLAAK 1997). The appearance of the Allerød soils in Poland is very similar to the Netherlands. The soils are slightly enriched in organic matter, homogenized by bioturbation and locally weakly podsolized (e. g. MANIKOWSKA 1991: 136). The Allerød soils described in eastern Germany on the other hand contain a clearly developed weathering horizon (Bw) and are classified as brown soils (Braunerde). The difference between the formation of brown soils and weakly podsolized soils can be climate related — the brown soil occurring in dryer climates — but since the weakly podsolized soils occur in both the Netherlands and Poland it is concluded that the formation of the brown soils is related to another soil forming factor. The brown soils described in eastern Germany by SCHLAAK were formed within the limits of the Weichselian glaciation while the slightly podsolized soils in the Netherlands and in Poland have been formed outside the limits of the Weichselian glaciation on a older substratum. The difference in age and mineralogical composition of the parent material from which the aeolian deposits were formed can explain the difference in soil formation. The Weichselian glacial deposits in eastern Germany were not strongly leached and weathered and consequently brown soils developed on the mineralogically richer aeolian material. In the Netherlands on the other hand the substratum had been intensively weathered during the Eemian and parts of the Weichselian and consequently slightly bleached and podsolized soils developed on the poorer aeolian material.

Younger Dryas (c. 11–10 ka)

During the Younger Dryas aeolian deposition started again in both western and central Europe (VAN DER HAMMEN 1951, 1971; VANDENBERGHE 1991; KASSE 1995; ISARIN et al. 1997; SCHLAAK 1997; NOWACZYK 1986; MANIKOWSKA 1991, 1994). The pollen diagram from Oud-Lutten (Fig. 3) shows that the peat was covered by sand after the Allerød-Younger Dryas transition, i. e. after 10,950 BP (HOEK 1997). The peat-sand contact, however, is sharp and there may be a hiatus between the end of peat formation and start of aeolian accumulation. Apparently the stabilizing vegetation cover of the Allerød period was destroyed. Formerly, the formation of the Younger Coversand II was attributed to the cold stadial conditions of the Younger Dryas period (VAN DER HAMMEN 1971; KOZARSKI & NOWACZYK 1991). However, recent results from the Netherlands demonstrate that the deposition of the so-called Younger Coversand II especially occurred in the second part of the Younger Dryas period, after c. 10.5 ka BP (see Tab. 1: phase III) (BOHNCKE et al. 1993; KASSE 1995; ISARIN et al. 1997). This means that aeolian deposition did not coincide completely with stadial conditions, since the first part of the Younger Dryas seems to have been colder than the second part (ISARIN & BOHNCKE, 1997). It is suggested that despite the coldness of the first part of the Younger Dryas aeolian activity was restricted by rather wet climatic conditions, while during the second part of the Younger Dryas, despite a slight climatic improvement, aeolian activity became more prominent due to an increased aridity.

The end of the Younger Coversand II deposition in the Netherlands is not well dated. In most cases the Younger Coversand II dune sands are overlain by the Holocene soil (mostly podsols). In a few cases peat that formed in dune deflation hollows has been dated (VAN GEEL et al. 1980/1981; SCHWAN 1991). The dates of 10,150 and 9,050 BP of the base of the peat show that the end of the Younger Coversand II deposition does not coincide necessarily with the end of the Younger Dryas biozone. Dune formation may have continued, perhaps locally, during the Preboreal period (KASSE et al. 1995). This continuation of aeolian deposition in the early Preboreal was also reported in Germany and Poland (SCHLAAK 1997; KOZARSKI 1990; MANIKOWSKA 1991, 1994). This means that the reconstruction of wind directions during the Younger Dryas based on the surface dune morphology (cf. ISARIN et al. 1997) may be hazardous. Although the sediment has been deposited during the Younger Dryas, the dune morphology may be a last-stage feature of Preboreal age.

In the Netherlands aeolian deposition during this period occurred in two regions:

i. higher places within the coversand area and
ii. along river valleys.

Higher well-drained locations of the previously deposited Older Coversand II and Younger Coversand I were reworked by the wind, but no new material was added to the sites. Especially on the higher interfluves between brooks and on coversand ridges this reworking is evident (VAN DER HAMMEN 1951). Apparently, not only the coldness of the Younger Dryas stadial, but also the increased aridity of the period caused the destruction of the vegetation cover and deflation of the surface.

The second region of large-scale Younger Dryas aeolian deposition is associated with river valleys. Along the large rivers like the Maas and Scheldt but also along smaller brooks extensive dune fields were formed on the east banks, indicating a westsouthwesterly wind during the Younger Dryas (KASSE 1995; ISARIN et al. 1997). The source of these aeolian sands is not by reworking of older coversands but from the Younger Dryas floodplains. Due to the Younger Dryas stadial conditions and related higher peak discharges the floodplain morphology had changed. In the case of the Maas river the channel morphology had changed from meandering during the Allerød to multichannel during the Younger Dryas. From these wider channel belts sand was deflated and accumulated along the floodplains.

From Germany and Poland it has been reported that the impact of the Younger Dryas on aeolian activity and dune formation was not so important. The already existing dune fields of the preceding Older Dryas period were slightly transformed but new dune fields were not formed (e. g. MANIKOWSKA 1991). The reason for this limited effect of the Younger Dryas cooling and aridity on the aeolian environment may be twofold. First, in contrast to the Netherlands, the vegetation decline and opening of the vegetation cover seems to have been less severe. The Younger Dryas vegetation remained dominated by forest elements (birch and pine) and bare ground susceptible to deflation did not occur frequently (BOHNCKE et al. 1995). Secondly, a change in river pattern from meandering to braided has not been reported frequently for the Polish rivers (KOZARSKI 1983; STARKEL 1995 and references cited there). Consequently, in contrast to the Netherlands, rivers maintained their meandering pattern and sediments were not easily deflated from the meandering channels and river-connected dune belts did not form.

4. Conclusions

Previously four coversand deposits have been distinguished each associated with a distinct cold period. In the foregoing it has been discussed that these four phases are not equally important and that their timing has to be adjusted. It is proposed here to define three phases of aeolian sand deposition in west and central Europe (Tab. 1):

Phase I (c. 20–14 ka BP): In this period of the Last Glacial Maximum aeolian processes have been very important, but the preservation of primary aeolian deposits has been low. The presence of continuous permafrost resulted in strong overland flow and aeolian sediments were reworked by water and deposited in valley floodplains as shallow fluvial deposits. Locally in the valley floodplains the reworked aeolian sediments alternate with primary aeolian beds and therefore have been described as fluvio-aeolian deposits. Formerly, they have been called the Older Coversand I. The sediments are especially extensive in floodplains but they normally lack outside the river plain. The sedimentary sequences from this period show that towards the end of this phase the fluvial activity decreased and aeolian deposition increased, which has been explained by an increase of the climatic aridity culminating in the formation of the Beuningen Gravel Bed.

Phase II (c. 14–11.9 ka): In this period following the Last Glacial Maximum the climate had already ameliorated, but deep seasonal frost still occurred. The aridity during this phase was strong, especially from c. 14 to 12.5 ka, since structures indicating surficial flow are nearly absent. Aeolian deposition occurred especially in sand sheets although locally in Poland low dunes were formed. Preservation of the aeolian sediments was strong and the deposits of this phase are the most important and widespread in western and central Europe. The sediments often have a characteristic alternating bedding of fine sandy and loamy fine sandy laminae in the lower part and horizontal bedding of fine sand in the upper part. Formerly, sediments from this phase were called the Older Coversand II and Younger Coversand I. Only locally the sediments from this phase can be subdivided in a phase II a and II b because of the presence of an organic layer of Bølling age. More often, however, Bølling age sediments are not present and the aeolian activity continued during the Bølling and Older Dryas periods until the start of the Allerød (phase II). In the Netherlands mostly sand sheets and low dunes were formed while further east sand sheets and high dunes were formed because of the stronger development of the vegetation cover in central Europe.
The Bølling period, especially from c. 12.5 ka onwards, is a period of decreased aeolian activity and increased organic production perhaps due to increased climatic humidity associated with the rapid Late Glacial climatic improvement. Local accumulations of silt

and organic material occurred in lower lying areas e. g. in front of coversand ridges. On higher ground, however, aeolian activity continued as indicated by the sandy character and sandy intercalations in organic deposits from this period.
Aeolian activity increased again in the later part of the Bølling (Tab. 1: phase II b) and culminated during the Older Dryas period. It is concluded that phase II b spans a longer time period from the late Bølling to 11.9 ka BP and is not strictly related to the short Older Dryas period as was previously stated for the Younger Coversand I. The Allerød period forms a gap in the aeolian record and landscape stability was complete by a closed forest cover. The short-lived cold Gerzensee oscillation did not lead to landscape instability and aeolian activity.

Phase III (c. 11–9 ka): The Younger Dryas stadial is reflected by renewed aeolian activity, but the onset of aeolian deposition is not always coinciding with the start of the Younger Dryas. It is concluded that the maximal aeolian activity occurred in the second part of the Younger Dryas after c. 10.5 ka. The Younger Dryas cooling was reflected almost immediately in the vegetation composition, but it is postulated that there was a time lag between climatic cooling and deterioration of the closed forest cover to such an extent that deflation could not occur. In addition to cooling also more intense aridity is held responsible for the opening of the vegetation cover. Phase III aeolian activity has been established in two different morphological regions. Firstly, high-lying, more drought-susceptible coversands of the previous phase II were remodelled or transformed during this phase into low dunes in the Netherlands, but new aeolian material was not supplied. Secondly, river dune fields were newly formed on the eastern banks of rivers, because of the change in channel geometry from meandering to braided at the start of the Younger Dryas.

Acknowledgments

ROEL VAN ELSAS and PETER WILLEKES are thanked for the pollen analysis; WIM HOEK and SJOERD BOHNCKE for preparing the pollen diagram and HENRY SION for drawing the figures.

5. References

BATEMAN, M. D. (1995): Thermoluminescence dating of the British coversand deposits. — Quaternary Science Reviews, **14**: 791–798.

BOHNCKE, S., KASSE, C. & VANDENBERGHE, J. (1995): Climate induced environmental changes during the Vistulian Lateglacial at Żabinko, Poland. — Quaestiones Geographicae, Special Issue, **4**: 43–64.

— VANDENBERGHE, J. & HUIJZER, A. S. (1993): Periglacial environments during the Weichselian Late Glacial in the Maas valley, the Netherlands. — Geologie en Mijnbouw, **72**: 193–210.

— WIJMSTRA, L., VAN DER WOUDE, J. & SOHL, H. (1988): The Late-Glacial infill of three lake successions in The Netherlands: Regional vegetational history in relation to NW European vegetational developments. — Boreas, **17**: 385–402.

BÖSE, M. (1991): A palaeoclimatic interpretation of frost-wedge casts and aeolian sand deposits in the lowlands between Rhine and Vistula in the Upper Pleniglacial and Late Glacial. — Z. Geomorph. N. F., Suppl.-Bd. **90**: 15–28; Berlin, Stuttgart.

BRUSSAARD. L. & RUNIA, L. T. (1984): Recent and ancient traces of scarab beetle activity in sandy soils of the Netherlands. — Geoderma, **34**: 229–250.

DE GANS, W. (1981): The Drentsche Aa valley system. — Thesis Vrije Universiteit Amsterdam: 132 p.; Amsterdam (Rodopi).

DIJKMANS, J. W. A. (1990): Niveo-aeolian sedimentation and resulting sedimentary structures; Søndre Strømflord area, Western Greenland. — Permafrost and Periglacial Processes, **1**: 83–96.

DÜCKER, A. & MAARLEVELD, G. C. (1957): Hoch- und spätglaziale äolische Sande in Nordwestdeutschland und in den Niederlanden. — Geol. Jahrb., **73**: 215–234.

GOOD, T. R. & BRYANT, I. D. (1985): Fluvio-aeolian sedimentation — an example from Banks Island, N. W. T., Canada. — Geografiska Annaler, **67A**: 33–46.

GOŹDZIK, J. (1991): Sedimentological record of aeolian processes from the Upper Plenivistulian and the turn of Pleni- and Late Vistulian in Central Poland. — Z. Geomorph. N. F., Suppl.-Bd. **90**: 51–60; Berlin, Stuttgart.

HIJSZELER, C. C. W. J. (1957): Late-Glacial human cultures in The Netherlands. — Geologie en Mijnbouw, **19**: 288–302.

HOEK, W. (1997): Palaeogeography of Lateglacial vegetations. Aspects of Lateglacial and Early Holocene vegetation, abiotic landscape, and climate in The Netherlands. — Thesis Vrije Universiteit Amsterdam: 147 p.; Utrecht (Drukkerij Elinkwijk b.v.).

HUISINK, M. (1998): Changing river styles in response to Weichselian climate changes in the eastern Netherlands. — Sedimentary Geology, submitted.

ISARIN, R. F. B. & BOHNCKE, S. J. P. (1998): Summer temperatures during the Younger Dryas in north-western and central Europe inferred from climate indicator plant species. — Quaternary Research, in press.

— RENSSEN, H. & Koster, E. A. (1997): Surface wind climate during the Younger Dryas in Europe as inferred from aeolian records and model simulations. — Palaeogeography, Palaeoclimatology, Palaeoecology, **134**: 127–148.

KASSE, C. (1995): Younger Dryas climatic changes and aeolian depositional environments. — In: TROELSTRA, S. R., VAN HINTE, J. E. & GANSSEN, G. M. [eds.]: The Younger Dryas. — Koninklijke Nederlandse Akademie van Wetenschappen Verhandelingen, Afd. Natuurkunde, Eerste Reeks, **44**: 27–31.

— (1997): Cold-climate sand-sheet formation in North-Western Europe (c. 14–12.4 ka); a response to permafrost degradation and increased aridity. — Permafrost and Periglacial Processes, **8**: 295–311.

— VANDENBERGHE, J. & BOHNCKE, S. (1995): Climatic change and fluvial dynamics of the Maas during the late Weichselian and early Holocene. — In: FRENZEL, B. [ed.]:

European river activity and climatic change during the Lateglacial and early Holocene. ESF Project European Palaeoclimate and Man, Special Issue 9. Paläoklimaforschung/Palaeoclimate Research, **14**: 123–150.

KOLSTRUP, E. (1980): Climate and stratigraphy in Northwestern Europe between 30.000 B.P. and 13.000 B.P., with special reference to The Netherlands. — Mededelingen Rijks Geologische Dienst, **32-15**: 181–253.

KOSTER, E. A. (1988): Ancient and modern cold-climate aeolian sand deposition: a review. — J. Quaternary Science, **3**: 69–83.

— & DIJKMANS, J. W. A. (1988): Niveo-aeolian deposits and denivation forms, with special reference to the Great Kobuk Sand Dunes, Northwestern Alaska. — Earth Surface Processes and Landforms, **13**: 153–170.

KOZARSKI, S. (1983): River channel changes in the middle reach of the Warta valley, Great Polish Lowland. — Quaternary Studies in Poland, **4**:159–169.

— (1990): Pleni and Late Vistulian aeolian phenomena in Poland: new occurrences, palaeoenvironmental and stratigraphic interpretations. — Acta Geographica Debrecina 1987–1988, **26–27**: 31-45.

— (1993): Late Plenivistulian deglaciation and the expansion of the periglacial zone in NW Poland. — Geologie en Mijnbouw, **72**: 143–157.

— & NOWACZYK, B. (1991): The late Quaternary climate and human impact on aeolian processes in Poland. — Z. Geomorph. N. F., Suppl.-Bd. **83**: 29–37; Berlin-Stuttgart.

MANIKOWSKA, B. (1991): Vistulian and Holocene aeolian activity, pedostratigraphy and relief evolution in Central Poland. — Z. Geomorph. N. F., Suppl.-Bd. **90**: 131–141; Berlin-Stuttgart.

— (1994): État des études des processus éoliens dans la région de Łódź (Pologne Centrale). — Biuletyn Peryglacjalny, **33**: 107–131.

MOL, J. (1997): Fluvial response to Weichselian climate changes in the Niederlausitz (Germany). — J. Quaternary Science, **12**: 43–60.

— & RHODES, E. J. (1997): Optical dating of quartz from Weichselian fluvial deposits in eastern Germany. — In: MOL, J. : Fluvial response to climate variations. The Last Glaciation in eastern Germany. — Thesis Vrije Universiteit Amsterdam; Enschede (Febodruk BV). — [Submitted to J. Quaternary Science].

— VANDENBERGHE, J., KASSE, K. & STEL, H. (1993): Periglacial microjointing and faulting in Weichselian fluvio-aeolian deposits. — J. Quaternary Science, **8**: 15–30.

NOWACZYK, B. (1986): The age of dunes, their textural and structural properties against atmospheric circulation pattern of Poland during the Late Vistulian and Holocene. — Adam Mickiewicz University Press, Seria Geografia, **28**: 245 p.

SCHLAAK, N. (1997): Äolische Dynamik im brandenburgischen Tiefland seit dem Weichselspätglazial. — Arbeitsberichte Geographisches Institut, Humboldt-Universität zu Berlin, **24**: 58 p.

SCHWAN, J. (1986): The origin of horizontal alternating bedding in Weichselian aeolian

sands in northwestern Europe. — Sedimentary Geology, **49**: 73–108.
— (1987): Sedimentologic characteristics of a fluvial to aeolian succession in Weichselian Talsand in the Emsland. — Sedimentary Geology, **52**: 273–298.
— (1991): Palaeowetness indicators in a Weichselian Late Glacial to Holocene aeolian succession in the southwestern Netherlands. — Z. Geomorph. N. F., Suppl.-Bd. **90**: 155–169; Berlin, Stuttgart.
— & VANDENBERGHE, J. (1991): Weichselian Late Pleniglacial fluvio-aeolian deposits and cryogenic structures. — Excursion guide Symposium periglacial environments in relation to climatic change, Maastricht/Amsterdam, 3–6 May 1991, Vrije Universiteit Amsterdam: 68–77; Amsterdam.
STAPERT, D. & VEENSTRA, H. J. (1988): The section at Usselo; brief description, grain-size distributions, and some remarks on the archaeology. — Palaeohistoria, **30**: 1–28.
STARKEL, L. (1995): The place of the Vistula river valley in the late Vistulian - early Holocene evolution of the European valleys. — In: FRENZEL, B. [ed.]: European river activity and climatic change during the Lateglacial and early Holocene. — ESF Project European Palaeoclimate and Man, Special Issue 9. Paläoklimaforschung/ Palaeoclimate Research, **14**: 75–88.
VAN DEN BERG, M. W. & SCHWAN, J. C. G. (1996): Millennial climatic cyclicity in Weichselian Late Pleniglacial to early Holocene fluvial deposits of the river Maas in the southern Netherlands. — In: VAN DEN BERG, M. W.: Fluvial sequences of the Maas: a 10 Ma record of neotectonics and climate change at various time-scales. — Thesis Agricultural University: 181 p.; Wageningen.
VANDENBERGHE, J. (1991): Changing conditions of aeolian sand deposition during the last deglaciation period. — Z. Geomorph. N. F., Suppl.-Bd. **90**: 193–207; Berlin, Stuttgart.
— & KROOK, L. (1981): Stratigraphy and genesis of Pleistocene deposits at Alphen (southern Netherlands). - Geologie en Mijnbouw, **60**: 417–426.
— & VAN HUISSTEDEN, J. (1988): Fluvio-aeolian interaction in a region of continuous permafrost. — Proceedings V International Conference on Permafrost, Trondheim, Norway: 876–881; Trondheim.
VAN DER HAMMEN, T. (1951): Late-Glacial flora and periglacial phenomena in The Netherlands. — Thesis State University Leiden: 183 p.; Leiden (Eduard Ijdo N.V.).
— (1971): The Upper Quaternary stratigraphy of the Dinkel valley. — In: VAN DER HAMMEN, T. & WIJMSTRA, T. A. [eds.]: The Upper Quaternary of the Dinkel valley (Twente, Eastern Overijssel, The Netherlands). — Mededelingen Rijks Geologische Dienst, **22**: 59–72.
— & WIJMSTRA, T. A. [eds.] (1971): The Upper Quaternary of the Dinkel valley (Twente, Eastern Overijssel, The Netherlands). — Mededelingen Rijks Geologische Dienst, **22**: 55–213.
VAN GEEL, B., BOHNCKE, S. J. P. & DEE, H. (1980/81): A palaeoecological study of an upper Late Glacial and Holocene sequence from "De Borchert", The Netherlands.

— Review of Palaeobotany and Palynology, **31**: 359–448.

— COOPE, G. R. & VAN DER HAMMEN, T. (1989): Palaeoecology and stratigraphy of the Lateglacial type section at Usselo (The Netherlands). — Review of Palaeobotany and Palynology, **60**: 25–129.

VAN HUISSTEDEN, J. (1990): Tundra rivers of the last glacial: sedimentation and geomorphological processes during the Middle Pleniglacial in Twente, Eastern Netherlands. — Mededelingen Rijks Geologische Dienst, **44**: 1–138

— VANDENBERGHE, J. & VAN GEEL, B. (1986): Late Pleistocene stratigraphy and fluvial history of the Dinkel Basin (Twente, Eastern Netherlands). — Eiszeitalter und Gegenwart, **36**: 43–59.

VINK, A. P. A. & SEVINK, J. (1971): Soils and paleosols in the Lutterzand. - In: VAN DER HAMMEN, T. & WIJMSTRA, T. A. [eds.]: The Upper Quaternary of the Dinkel valley (Twente, Eastern Overijssel, The Netherlands). — Mededelingen Rijks Geologische Dienst, **22**: 165–185.

WOO, M.-K. & WINTER, T. C. (1993): The role of permafrost and seasonal frost in the hydrology of northern wetlands in North America. — J. of Hydrology, **141**: 5–31.

ZEEBERG, J. J. (1998): The European sand belt in eastern Europe — an comparison of Late Glacial dune orientation with GCM simulation results. — Boreas, **27**: 127–139.

Address of the author:

Dr. CORNELIS KASSE, The Netherlands Centre for Geo-ecological Research, Faculty of Earth Sciences, Vrije Universiteit, De Boelelaan 1085, NL-1081 HV Amsterdam, e-mail: kask@geo.vu.nl